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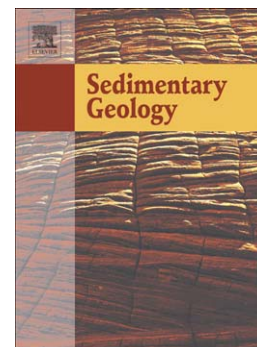
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Continental growth seen through the sedimentary record

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Abstract

Sedimentary rocks and detrital minerals sample large areas of the continental crust, and they are increasingly seen as a reliable archive for its global evolution. This study presents two approaches to model the growth of the continental crust through the sedimentary archive. The first builds on the variations in U-Pb, Hf and O isotopes in global databases of detrital zircons. We show that uncertainty in the Hf isotope composition of the mantle reservoir from which new crust separated, in the $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of that new crust, and in the contribution in the databases of zircons that experienced ancient Pb loss(es), adds some uncertainty to the individual Hf model ages, but not to the overall shape of the calculated continental growth curves. The second approach is based on the variation of Nd isotopes in 645 worldwide fine-grained continental sedimentary rocks with different deposition ages, which requires a correction of the bias induced by preferential erosion of younger rocks through an erosion parameter referred to as K . This dimensionless parameter relates the proportions of younger to older source rocks in the sediment, to the proportions of younger to older source rocks present

in the crust from which the sediment derived. We suggest that a Hadean/Archean value of $K = 1$ (i.e., no preferential erosion), and that post-Archean values of $K = 4-6$, may be reasonable for the global Earth system. Models built on the detrital zircon and the fine-grained sediment records independently suggest that at least 65% of the present volume of continental crust was established by 3 Ga. The continental crust has been generated continuously, but with a marked decrease in the growth rate at ~ 3 Ga. The period from >4 Ga to ~ 3 Ga is characterised by relatively high net rates of continental growth ($2.9-3.4 \text{ km}^3 \text{ yr}^{-1}$ on average), which are similar to the rates at which new crust is generated (and destroyed) at the present time. Net growth rates are much lower since 3 Ga ($0.6-0.9 \text{ km}^3 \text{ yr}^{-1}$ on average), which can be attributed to higher rates of destruction of continental crust. The change in slope in the continental growth curve at ~ 3 Ga is taken to indicate a global change in the way bulk crust was generated and preserved, and this change has been linked to the onset of subduction-driven plate tectonics. At least 100% of the present volume of the continental crust has been destroyed and recycled back into the mantle since ~ 3 Ga, and this time marks a transition in the average composition of new continental crust. Continental crust generated before 3 Ga was on average mafic, dense, relatively thin (<20 km) and therefore different from the calc-alkaline andesitic crust that dominates the continental record today. Continental crust that formed after 3 Ga gradually became more intermediate in composition, buoyant and thicker. The increase in crustal thickness is accompanied by increasing rates of crustal reworking and increasing input of sediment to the ocean. These changes may have been accommodated by a change in lithospheric strength at around 3 Ga, as it became strong enough to support high-relief crust. This time period therefore indicates when significant volumes of continental crust started to become emergent and were available for erosion and weathering, thus impacting on the composition of the atmosphere and the oceans.

Keywords

Continental growth; Hadean/Archaean; plate tectonics; zircon; shale; U-Pb/Hf/Nd/O isotopes

1. Introduction

The continental crust has evolved over billions of years, helping to create the environment we live in and the resources we depend on. Understanding how and when it formed is an important step in unravelling the evolution of the Earth system, yet these questions remain matters of considerable discussion. This is because most rocks in the geological record derive from pre-existing crustal rocks (e.g., Hutton, 1788), and so it remains difficult to unpick from the present record processes of crust generation from processes of crustal destruction, reworking and preservation (Hawkesworth et al., 2009, 2010; Belousova et al., 2010; Condie et al., 2011; Voice et al., 2011; Dhuime et al., 2012; Arndt, 2013; Vervoort and Kemp, 2016; Iizuka et al., 2017). The present-day composition of the bulk continental crust is well constrained, with studies converging towards an average andesitic composition (i.e., SiO_2 ~57–65%) (Taylor, 1964; Rudnick and Gao, 2003; Hacker et al., 2011). Its intermediate/felsic composition is widely explained by a two-stage model in which juvenile (proto)continental crust is extracted from the mantle, followed by differentiation through partial melting and/or fractional crystallization (e.g., Rudnick, 1995). The rates and timing of the net addition of new crust to the continental landmass, commonly referred to as 'continental growth', remain matters of debate.

A number of studies have attempted to evaluate the volumes of continental crust through the evolution of the Earth, using different approaches and proxies. A summary of cumulative growth curves, which are all anchored by the present volume of the continental crust at 100%,

is presented in Fig. 1. The most extreme scenarios suggest that the volume of continental crust at the Archaean-Proterozoic boundary may have been less than 25%, or more than 100%, of the present-day volume (Fig. 1, curves 'V&J' and 'F', respectively). Some curves are relatively smooth (e.g., curves 'O'N' and 'Be'), and so represent continuous growth of the continental crust, whereas others are stepped (e.g., curves 'T&M' and 'C&A') suggesting episodic growth. Between such end-members a number of scenarios have been envisaged, where the locus and the intensity of breaks in slope in the curves are linked to changes in the timing and rates of continental growth.

Sediments derived from continental sources (here after referred to as 'continental sediments') cover vast tracks of continental crust, with for instance the 15 largest rivers draining ~30% of the surface area of the continental crust (Goldstein et al., 1984; Milliman and Syvitski, 1992). Ancient sediments deposited million to billions of years ago preserve information on the continental crust that has been destroyed and is no longer available for sampling through magmatic rocks (McLennan and Taylor, 1982; O'Nions et al., 1983; Allègre and Rousseau, 1984; Goldstein and Arndt, 1988; Goldstein and Jacobsen, 1988). As a consequence, the sedimentary record is increasingly seen as a reliable archive for the global evolution of the continental crust through time, and over the last decade there has been increasing interest in using the information available in the ever-growing databases of detrital zircons to model the growth of the continental crust (Rino et al., 2004; Condie et al., 2005; Wang et al., 2009, 2011; Belousova et al., 2010; Iizuka et al., 2010; Condie et al., 2011; Dhuime et al., 2012; Parman, 2015). In parallel there has been a recent interest in exploring the processes that bias the sedimentary record (Dhuime et al., 2011b; Cawood et al., 2013) in order to better constrain growth models based on bulk sediment data (Allègre and Rousseau, 1984; Michard et al., 1985; Dia et al., 1990a).

The mineral zircon is a key minor constituent of the igneous rocks generated in the production of the continental crust, which together with its physiochemical resilience and the development of microanalytical techniques enabling the rapid determination of its isotopic and trace element compositions, makes zircon an important archive of the evolution of the continental crust. Importantly, detrital zircons remain one of the few records of geological processes in the first 500 million years of Earth's history, a period for which no rocks are known to have been preserved (Mojzsis et al., 2001; Wilde et al., 2001; Watson and Harrison, 2005; Blichert-Toft and Albarède, 2008; Harrison et al., 2008; Hopkins et al., 2008, 2010; Harrison, 2009; Kemp et al., 2010; Trail et al., 2011; Bell et al., 2015a). Initial studies of the continental record using zircon relied on their U-Pb age distribution in modern river sediments to model continental growth (Rino et al., 2004, 2008). The development of *in situ* Lu-Hf analyses in zircon by laser ablation multi-collector inductively coupled plasma mass spectrometry (LA-MC-ICP-MS) (Griffin et al., 2000; Iizuka et al., 2005; Kemp et al., 2005), rapidly followed by an explosion of the number of LA-MC-ICP-MS laboratories, has highlighted that 80–90% of zircons analysed have Hf isotopic compositions that are not in equilibrium with the composition of new continental crust extracted from the upper mantle at the time they crystallised (Belousova et al., 2010; Voice et al., 2011; Roberts and Spencer, 2015). This reaffirms that most rocks of the continental crust derive—at least in part—from pre-existing crustal material, as was argued by Hutton in the eighteenth century (Hutton, 1788). As a consequence, continental growth curves based solely on distributions of U-Pb crystallisation ages will be strongly influenced by ages that reflect the reworking of continental crust, and as such they can only be considered as reflecting the *minimum* volume of continental crust that was present at a particular time (Komiya, 2011; Hawkesworth et al., 2013).

An initial difficulty in modelling the growth of the continental crust is how to evaluate the proportion of newly generated crust over that of reworked pre-existing crust at different times, as recorded in zircons and continental sediments – and the extent to which this proportion has changed through time (e.g., Belousova et al., 2010; Iizuka et al., 2010; Dhuime et al., 2012). In this contribution we discuss how variations in U-Pb, Hf and O isotopes in detrital zircons can be used to model the growth of the continental crust, and the strengths and limitations of this approach. We demonstrate through the combination of this approach, and an independent approach based on the variation of Nd isotopes in fine-grained continental sedimentary rocks with different deposition ages, building on the equations of Allègre and Rousseau (1984), that by 3 Ga the volume of the continental crust was at least 65% of the present volume. Finally we use these results, along with data from the literature, to develop a preliminary model for the evolution of the continental crust that takes into account changes in the rates at which crust was generated and destroyed.

2. Detrital zircon record

2.1. Continental growth models based on the combination of U-Pb, Hf (and O) isotopes

Belousova et al. (2010) were the first to model the growth of the continental crust on the basis of variations in the proportion of new relative to reworked crust, using a worldwide U-Pb and Hf isotopes database of over 13,000 zircons, largely of detrital origin. Their database was broken down into 45 time slices, each of 100 Ma duration. For each 100 Ma interval, the proportion of juvenile crust addition (X_{juv}) was calculated by counting the number of zircons with U-Pb ages ($N_{U-Pb\ age}$) in that interval, and the number of zircons with Hf model ages ($N_{model\ age}$) that range within the limits of the same period, using the following equation:

$$X_{\text{juv}} = N_{\text{model age}} / (N_{\text{U-Pb age}} + N_{\text{model age}}). \quad (1)$$

A continental growth curve was built from the cumulative proportions of juvenile crust addition through time (Fig. 2, black curve), and this model suggests that ~50% of the present volume of continental crust was present by 3 Ga (Belousova et al., 2010).

However, the Hf isotope composition of the zircons analysed may reflect mixtures of both juvenile and older reworked material (Kemp et al., 2007), and so individual model ages may be hybrid ages and hence not represent true periods of crust formation (Arndt and Goldstein, 1987). Thus a significant uncertainty remains over the shape of continental growth curves that rely solely on model ages. This issue can in part be tackled by combining oxygen isotopes with U-Pb and Hf isotopes (Hawkesworth and Kemp, 2006b; Kemp et al., 2006). 'Mantle-like' zircons, i.e., zircons in high temperature equilibrium with mantle-derived magmas, have a narrow range of $\delta^{18}\text{O}$ (typically $\delta^{18}\text{O} = 5.3 \pm 0.6\text{‰}$ (2 s.d.; Valley et al., 1998)), and $\delta^{18}\text{O}$ in zircons are higher or lower when their parent magmas contain a contribution of supracrustal material (e.g., sediments deriving from pre-existing crust). Periods of juvenile crust formation are taken to be characterised by zircons that have mantle-like $\delta^{18}\text{O}$ and limited variation in their Hf model ages irrespective of their U-Pb ages (Kemp et al., 2006). Conversely periods dominated by crustal reworking produce 'supracrustal' zircons, typically with elevated $\delta^{18}\text{O}$ values (Valley et al., 2005; Spencer et al., 2014) and a large variation of hybrid Hf model ages at similar U-Pb ages (Kemp et al., 2006; Vervoort and Kemp, 2016). To the extent that Hf isotope ratios of supracrustal zircons may represent mixtures, they will not record true periods of crustal growth (Kemp et al., 2006).

A difficulty in developing models of continental growth based on zircons is therefore in evaluating the proportions of new crust formation ages and hybrid ages in large datasets

representative of the average continental crust, and for which O isotope data are not available (Belousova et al., 2010; Condie et al., 2011; Voice et al., 2011; Roberts and Spencer, 2015). Dhuime et al. (2012) explored the extent to which the variations between Hf model ages and $\delta^{18}\text{O}$ in zircons for which such data are available might be parameterised in order to better correct for the bias induced by the presence of zircons with hybrid Hf model ages in datasets with no O isotope data. They found, using a global U-Pb, Hf and O isotope database of 1376 zircons analyses, that the proportion of 'true' crust formation to hybrid Hf model ages calculated for every 100 Ma time slice is ~ 0.73 between the Hadean and 3.2 Ga. After 3.2 Ga this proportion gradually decreases to ~ 0.2 at 2 Ga, then increases to ~ 1 at present, following a second-order polynomial relationship with an $R^2 = 0.91$ (Fig. 2, inset). Dhuime et al. (2012) used this relationship to correct for the contribution of hybrid Hf model ages in a worldwide database of 6972 U-Pb and Hf analyses of zircons from young sediments. They defined the proportion of juvenile crust addition (X_{juv}) for every 100 Ma time slice of this database as:

$$X_{\text{juv}} = N_{\text{NC ages}} / (N_{\text{NC ages}} + N_{\text{RC ages}}), \quad (2)$$

with $N_{\text{NC ages}}$ = number of calculated new crust ages, as:

$$N_{\text{NC ages}} = N_{\text{model ages}} * 0.73 \text{ t for } t_{>3.2 \text{ Ga}}, \quad (3)$$

$$N_{\text{NC ages}} = N_{\text{model ages}} (2.894\text{E-}7 t^2 + 1.085\text{E-}3 t + 1.243) \text{ for } t_{0-3.2 \text{ Ga}}; \quad (4)$$

and $N_{\text{RC ages}}$ = number of reworked crust ages, calculated as the difference between the total number of U-Pb ages within a given time slice, and the number of zircons with similar U-Pb ages and Hf model ages within the same time slice.

Dhuime et al. (2012) calculated the continental growth curve presented in Fig. 2, from the cumulative proportions of juvenile crust addition through time. This curve suggests that by 3 Ga the volume of the continental crust was $\sim 65\%$ of the present volume, and that there was a decrease in the net rate of growth at that time. These features are not observed in growth

curves calculated from large U-Pb and Hf in zircon databases in the absence of O isotope data (Belousova et al., 2010; Roberts and Spencer, 2015).

2.2. Critical parameters influencing zircon-based continental growth models

Continental growth models that rely on zircon Hf model ages are regarded as controversial (Arndt, 2013; Arndt and Davaille, 2013; Guitreau and Blichert-Toft, 2014; Roberts and Spencer, 2015; Couzinié et al., 2016; Payne et al., 2016; Vervoort and Kemp, 2016; Iizuka et al., 2017; Rollinson, 2017), principally because the calculation of zircon Hf model ages may be influenced by a number of factors; (i) the Hf isotope composition of the reference reservoir from which model ages are calculated; (ii) the $^{176}\text{Lu}/^{177}\text{Hf}$ isotope ratio of the crustal source of the magmas from which the zircons crystallised; and (iii) underestimation of the crystallisation age of apparently concordant zircons that experienced ancient Pb loss(es). The extent to which zircon datasets preserve a representative record of the evolution of the crust may also play a role in continental growth models. The influence of these parameters on the shape of the global continental growth curves is evaluated below.

2.2.1. Influence of the isotope composition of the source reservoir of the new continental crust

Geochemical mass balance calculations indicate that about 80% of the continental crust that is preserved today was generated along destructive plate margins (Rudnick, 1995; Hawkesworth and Kemp, 2006a). This implies that subduction-related magmas should be used to constrain the isotope composition of the continental crust at the time of its formation, at least for the period of time when plate tectonics has been operating. Consequently, Dhuime et al. (2011a) suggested that model ages, traditionally calculated for crust derived from the depleted mantle (e.g., DePaolo, 1981; Goldstein et al., 1984; DePaolo et al., 1991), should be calculated using the isotope composition of the new continental crust generated in subduction settings. The

present-day isotope composition of that new crust ($\epsilon_{\text{Hf}} = 13.2 \pm 1.1$) was determined from the data of modern intraoceanic magmatic arcs worldwide, and Dhuime et al. (2011a) proposed a linear evolution for the new crust, with a chondritic uniform reservoir (CHUR)-like isotope composition at the time of Earth's formation (Fig. 3). Model ages calculated from the Hf isotope composition of the new crust are up to 300 Ma younger than model ages calculated from the depleted mantle, and all calculations for the Dhuime et al. (2012) model presented in Fig. 2 were done using the Dhuime et al. (2011a) new crust evolution model.

In order to evaluate the extent to which the isotope composition of the source reservoir of the new continental crust may affect the shape of the continental growth curves calculated from the zircon data, continental growth curves were recalculated assuming two different models for the inferred source reservoir of the new continental crust: (i) a traditional depleted mantle-like evolution, from a CHUR-like isotope composition at the time of Earth's formation until $\epsilon_{\text{Hf}} = 17$ at present (Salters and Stracke, 2004; Workman and Hart, 2005) (Fig. 3A, green curve); and (ii) a CHUR-like evolution for the new crust between Earth's formation and ~ 3.8 Ga, as suggested by Vervoort et al. (2013), and followed by a linear evolution up to $\epsilon_{\text{Hf}} = 13.2$ at present (Fig. 3A, orange curve). The similar shapes of the continental growth curves obtained with these different methods (Fig. 3B, C) and the model of Dhuime et al. (2012) (Fig. 2) suggest that no significant bias results from the choice of the reference curve for the Hf isotope ratios of new continental crust in the calculation of model ages when large datasets are considered.

2.2.2. Influence of the $^{176}\text{Lu}/^{177}\text{Hf}$ ratio inferred for the crustal source

There is some uncertainty over the $^{176}\text{Lu}/^{177}\text{Hf}$ ratio of the crustal source of the magmas from which zircons crystallised, and whether this ratio may have changed through time (Pietranik

et al., 2008; Harrison, 2009; Kemp et al., 2010; Guitreau et al., 2012; Payne et al., 2016; Vervoort and Kemp, 2016). This uncertainty propagates into the calculation of Hf model ages, which makes the robustness of individual model ages difficult to evaluate. A value of $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$, typical for the bulk continental crust (e.g., Taylor and McLennan, 1995; Rudnick and Gao, 2003; Hacker et al., 2011), was preferred in the models of Belousova et al. (2010) and Dhuime et al. (2012) (Fig. 2), and a graphical representation of the calculation of the Hf model ages with this method is shown in Fig. 4A (grey segment and grey star (0)). In order to test the effect of having different $^{176}\text{Lu}/^{177}\text{Hf}$ ratios, two extreme case scenarios are explored: (i) a crustal source that is more felsic/evolved than the bulk crust ($^{176}\text{Lu}/^{177}\text{Hf} = 0.009$, red segment), and for which calculated model ages are represented by the red star (1) in Fig. 4A; and (ii) a crustal source that is more mafic than the bulk crust ($^{176}\text{Lu}/^{177}\text{Hf} = 0.022$, green segment), and for which calculated model ages are represented by the green star (2). The resulting continental growth curves are shown in Fig. 4B and 4C for $^{176}\text{Lu}/^{177}\text{Hf} = 0.009$ and $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$, respectively. The shapes of the continental growth curves in Figs. 2 and 4B,C are similar, which suggests that variations in the $^{176}\text{Lu}/^{177}\text{Hf}$ of the crustal sources have minimal impact in the calculation of continental growth models based on large sets of zircon data.

2.2.3. Influence of ancient Pb loss(es) on the calculation of crust formation ages

The calculation of zircon Hf model ages of crust formation critically relies on good estimates of the U-Pb crystallisation age of the zircon, as well as on the Hf isotope ratio of the zircon at the time of crystallisation. Greater confidence in the accuracy of the crystallisation ages can in principle be achieved by filtering discordant U-Pb analyses (e.g., analyses with a discordancy greater than 5–10%) in zircon databases, however some uncertainty remains over the crystallisation ages of detrital zircons with ancient (typically Archaean) concordant U-Pb ages

if the zircons have experienced ancient Pb loss(es). The effect of ancient Pb losses in the calculation of initial Hf isotope ratios and model ages of crust formation has been discussed in a number of studies (e.g., Kemp and Whitehouse, 2010; Guitreau and Blichert-Toft, 2014; Claesson et al., 2015; Vervoort and Kemp, 2016), and a graphical summary is given in Fig. 5A. The proportion of detrital zircons that have experienced ancient Pb losses remains difficult to address in large datasets, and the extent to which analyses affected by Pb loss(es) may affect the shape of continental growth curve remains unclear. One simple test is to select grains with Archaean ages in large U-Pb and Hf detrital zircons databases, then subtract a random amount of time from their crystallisation ages to simulate Pb loss, and use the altered/modified database to recalculate a continental growth curve. The Dhuime et al. (2012) dataset was selected, random amounts of time between 0–500 Ma were subtracted from the crystallisation ages of Archaean grains using an in-house Excel spreadsheet (the average reduction was ~250 Ma), and a new continental growth curve was calculated from the altered database (Fig. 5B). Its aspect remains similar to that of the curve obtained from the original dataset, which suggests that the effect of ancient Pb losses is small on the shape of continental growth curves based on large U-Pb and Hf datasets.

2.2.4. Influence of the nature and the size of zircon databases on continental growth models

The sampling strategy, and more specifically the extent to which specific sets of data best record the global evolution of the continental crust, is a key issue in continental growth studies. Belousova et al. (2010) used a database of over 13,000 zircons sampled worldwide. This database contains zircons of both magmatic and detrital origin, but it is dominated by detrital zircons, both from ancient and younger sediments. Dhuime et al. (2012) used a smaller database of ~7000 zircons for their model, containing only detrital zircons with deposition ages ranging from the late Palaeozoic to the present. Their choice was driven by

the assumption that since 'young' sediments typically contain zircons with a wide range of ages, they may provide records that are more representative of the entire magmatic history of the crust than zircons in igneous rocks or in old sediments. More recently Roberts and Spencer (2015) compiled the U-Pb and Hf data for magmatic and detrital zircons available in the literature ultimately to add nearly 30,000 analyses to the Belousova et al. (2010) database. In order to test the sensitivity of the Dhuime et al. (2012) continental growth model to the use of other databases, the equations of Dhuime et al. (2012) (see Section 2.1) were applied to both the Belousova et al. (2010) and the Roberts and Spencer (2015) databases to build the continental growth curves presented in Fig. 6A and 6B, respectively. These curves point towards slightly lower volume of ancient crust, as for instance the volume of >3 Ga crust is ~60% of the present volume in Fig. 6A and 6B, compared with ~65% in Fig. 2. Despite these differences, these curves can be considered as similar, especially in comparison with the range of growth curves in Fig. 1. Lower volumes of Archaean crust obtained from the Belousova et al. (2010) and the Roberts and Spencer (2015) databases may also reflect an underestimation of the proportions of juvenile crust at that time, which can be predicted in datasets that include ancient rocks and sediments (see equation (1)).

2.2.5. *Summary*

Difficulties in establishing the initial Hf isotope ratios of new crust, the $^{176}\text{Lu}/^{177}\text{Hf}$ of that new crust, and in the numbers of zircons that experienced ancient Pb loss(es) in the databases, adds some uncertainty to the individual model ages, but not to the overall shape of the continental growth curves being modelled (Figs. 2–5). The shape varies slightly depending on which U-Pb and Hf sets of data are used in the calculations (Fig. 6), but overall the conservation of the global aspect of the continental growth curves is reassuring as it strengthens approaches based on the combination of U-Pb ages and Hf model ages in large zircon databases.

3. Fine-grained sediments record

3.1. Background

Fine-grained continental sedimentary rocks, most of which are described as 'shales' in the literature, have been widely used to unravel the processes of crustal differentiation, maturation and reworking that occurred on large spatial and temporal scales (Nance and Taylor, 1976, 1977; McLennan and Taylor, 1982; Hamilton et al., 1983; O'Nions et al., 1983; Frost and O'Nions, 1984; Davies et al., 1985; Michard et al., 1985; Miller and O'Nions, 1985; Taylor and McLennan, 1991; Condie, 1993; Gao and Wedepohl, 1995; Jahn and Condie, 1995; Rudnick and Gao, 2003; Bindeman et al., 2016; Tang et al., 2016). The Rare Earth Elements (REE) contents of sediments have long been of particular interest (e.g., Nance and Taylor, 1976, 1977), because REE have a low solubility in water and so they are less subject to fractionation in the sedimentary cycle than many other elements. Sm and Nd are especially relevant because they are the parent and daughter elements of a long-lived radioactive decay chain, and because the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios show a limited range of variation in fine-grained continental sedimentary rocks (e.g., mean $^{147}\text{Sm}/^{144}\text{Nd} = 0.116 \pm 0.013$ s.d., from a compilation of 645 samples worldwide; references are provided in Section 3.3).

Allègre and Rousseau (1984) were the first to use the geochemical properties of the Sm-Nd system in shales to develop a model for the growth of the continental crust. More specifically, they used a box-model approach and the variation of Nd model ages in eight samples of Australian shales with deposition ages ranging 3.3–0.2 Ga to generate a continental growth curve. In their model the present-day continental crust was arbitrarily subdivided into seven segments with mean ages of 3.50, 2.75, 2.25, 1.75, 1.25, 0.75 and 0.25 Ga. These segments

were taken as the sources from which continental sediments were eroded, and then deposited every 500 Ma at $t_1 = 2.5$ Ga, $t_2 = 2$ Ga, $t_3 = 1.5$ Ga, $t_4 = 1$ Ga, $t_5 = 0.5$ Ga and $t_6 = 0$ Ga. These deposition ages constitute six consecutive steps in the box-model of Allègre and Rousseau (1984). For each step t_n , they calculated the mass fraction of juvenile crust present in the bulk sediment. To do so they considered that the sources of the sediments can be considered in terms of two continental blocks (Fig. 7): a young block of mass fraction $[x]$ made of juvenile crust (green block), and an older block of mass fraction $[1-x]$ that represents the average of all the continental segments formed in previous events (brown block). A major difficulty was to evaluate the extent to which the proportions of different source rocks in the sediments may be biased through erosion processes. Allègre and Rousseau (1984) argued that young orogens have a higher relief than older continental segments, and that since rocks are more prone to erosion as the relief increases due to their greater potential energy (Ahnert, 1970; Pinet and Souriau, 1988; Summerfield, 1991; Milliman and Syvitski, 1992; Summerfield and Hulton, 1994), the preferential erosion of high-relief young crust over low-relief old crust ultimately results in an over-representation of younger source rocks in continental sediments. It follows that in the two-block model of Allègre and Rousseau (1984), sediment extracted from the young block has a mass fraction, here termed $[y]$, such that $[y]$ is $> [x]$, because relatively more sediment is derived from the younger source block of mass fraction $[x]$. Reciprocally, sediment derived from the old block has a mass fraction $[1-y]$ that is $< [1-x]$ (Fig. 7).

The link between the relative proportions of source rocks of different ages in the catchment area, and the proportion of those source rocks present in the sediment analysed, can be expressed through an erosion parameter K . This parameter is dimensionless and was defined by Allègre and Rousseau (1984) as:

$$K = ([y]/[1-y]) / ([x]/[1-x]) \quad (5)$$

Using this parameterisation, Allègre and Rousseau (1984) calculated the mass fraction $[x]$ of the juvenile crust, for every step t_n of their box-model, using the following equation:

$$[x] = 1 / ((K \cdot [1-y]/[y]) + 1) \quad (6)$$

The mass fractions $[y]$ and $[1-y]$ in the sediment issued from the young block and the old block, respectively, were calculated from the mean age of the sediment T_{sediment} , for every step t_n of the box-model, using the following mixing equation:

$$T_{\text{sediment}} = [y] \cdot T_{\text{young}} + [1-y] \cdot T_{\text{old}} \quad (7)$$

with T_{sediment} = Nd model age of the sediment; T_{young} = age of the young block; and T_{old} = mean age of the old block made of all the continental segments formed in previous events.

By rearranging equation (7), the mass fraction $[y]$ becomes:

$$[y] = (T_{\text{sediment}} - T_{\text{old}}) / (T_{\text{young}} - T_{\text{old}}) \quad (8)$$

Once the value of $[y]$ is reinserted into equation (6), the only unknown parameter left to calculate $[x]$ is the value of erosion parameter K . Allègre and Rousseau (1984) used assumed values of $K = 2, 4$ or 6 in their model to generate three curves (Fig. 8A, red curves). The green curve in Fig. 8A is the growth curve for $K = 1$, i.e., when $[y] = [x]$ and no correction for the preferential erosion of the young crust over the older crust is applied to the model. The grey curves represent the effect of changing values of K , up to $K = 50$, on the global shape of the modelled growth curves. The higher the value of K , the more young crust is overrepresented in the sediment analysed, hence the larger the underestimation of the amount of old crust in the source of the sediment, and so the larger the volume of ancient continental crust that is actually present. These curves were established from the relative proportions of the mass fractions $[x]$ of juvenile crust in the segments of ages 2.75, 2.25, 1.75, 1.25, 0.75 and 0.25 Ga at the last step $t_6 = 0$ Ga of the box-model, i.e., in the present-day crust.

Thus the approach of Allègre and Rousseau (1984) seeks to evaluate the proportion of crustal segments of different formation ages in the crust at the present day, and then infers that they can be used as a proxy for the volumes of crust present at different times in the history of the crust. This is similar to the approach of Condie (1998), and more recently Condie and Aster (2010), who used the variation in U-Pb ages distribution of magmatic rocks with juvenile Nd isotope ratios as a proxy for the variation of crustal volumes through time (see Fig.1, curve 'C&A'). On that basis the Allègre and Rousseau (1984) model suggests that both the volume of >3 Ga juvenile rocks in the present crust and the volume of continental crust established by 3 Ga were ~25% of the present volume of crust for $K = 2$, and they increase to ~37% and ~46% for $K = 4$ and $K = 6$, respectively. These results contrast with the model proposed by Dhuime et al. (2012) from the zircon record (Fig. 2), in which the volume of continental crust present by 3 Ga is ~65% of the present volume. In the following sections we explore ways in which the Allègre and Rousseau (1984) model might be reconciled with models in which large volumes of crust were present early in Earth's history (e.g., Fig. 1). These include (i) variations in the value of the erosion parameter K ; (ii) extension of the initial Nd isotope data of Allègre and Rousseau (1984) on eight Australian shales to a worldwide dataset of 645 fine-grained continental sedimentary rocks considered to be more representative of the continental crust, and (iii) other ways in which continental growth models might be developed from the variations in Nd isotopes in fine-grained sediments.

3.2. Uncertainty over the global value of the erosion parameter K

Attempts to calculate the volume of crustal segments of different model ages from the Nd isotope ratios of continental sediments critically depend on the value of the erosion parameter K (Fig. 8). The value of this parameter that should be considered as representative of the global Earth system has been difficult to establish (Allègre and Rousseau, 1984; Jacobsen,

1988; Kramers and Tolstikhin, 1997; Nägler and Kramers, 1998; Kramers, 2002; Tolstikhin and Kramers, 2008), in part because only a few attempts have been made in measuring K in natural systems. Goldstein and Jacobson (1988) estimated a value of $K = 2.3$ from the average Nd isotope data of suspended loads in 31 rivers in North America, and they suggested that a value of $K = 2-3$ may have been applicable through much of Earth's history. Allègre et al. (1996) used the Nd isotope data of suspended loads in four rivers from the Amazon Basin, and the relative proportion of high-relief rocks in the rivers' catchment areas, to calculate a value of $K = 5.8$ for the Amazon Basin as a whole.

More recently Dhuime et al. (2011b) targeted the Frankland River in southwest Australia, a 'simple' system that only drains two continental blocks with distinctive age components (Cawood et al., 2003), the Archaean Yilgarn craton and the Archaean to Proterozoic Albany-Fraser orogen, in order to measure K *in situ* in the same way as this parameter was originally defined in the two-block model of Allègre and Rousseau (1984) (see Fig. 7). Dhuime et al. (2011b) calculated K in three sample sites across the Frankland river, located at ~120 km, ~60 km and 0 km, respectively from the river mouth (Fig. 9). They used three different approaches to determine the relative proportions of Yilgarn and Albany Fraser detritus in the sediment samples: (i) the distribution of U-Pb ages in zircons, (ii) the distribution of Hf model ages in zircons, and (iii) the Nd isotope composition of the bulk river sediments. They demonstrated that these three approaches give similar results, and they used these data and the proportions of the source rocks in the river catchment determined from a geological map to calculate K . They showed that K varies along the river, with values of 4–6, 9–10 and 15–17 at 120 km, 60 km and 0 km from the river mouth, respectively (Fig. 9). Importantly the values of K increase with the gradient of the river's profile, with the two samples providing the highest K values (i.e., 9–10 and 15–17) located below an inflection in the river's profile. This inflection reflects

an escarpment at ~80 km from the coast associated with Miocene-Pliocene uplift (Cawood et al., 2003), and so from this period the increase in gradient of the river's profile has resulted in preferential erosion of material from the younger block (i.e., the Albany-Fraser Orogen), and hence in higher calculated K values. Dhuime et al. (2011b) suggested that in order to avoid local perturbations, the stable segment of the Frankland River is best sampled above the escarpment, and that K values of 4–6 may be considered as better representative of mature river systems that sample large areas of continental crust.

Cawood et al. (2013) used the variation in the running median of the Hf isotope composition of zircons through time (Belousova et al., 2010) to argue that crustal melting is an important process in the continental record. They highlighted that crustal melting is often associated with crustal thickening, and hence with areas of high relief. As a consequence they suggested that the bias in the sedimentary record might be dominated by erosion in areas of high relief and, if correct, they suggested that the highest values of K so far measured in natural systems (i.e., $K = 15$ –17 in the Frankland River, see Fig. 9) may better account for the processes of erosion and deposition of the sediments that dominate the geological record. It is striking that for a value of $K = 15$, the continental growth curve of Allègre and Rousseau (1984) (Fig. 8A) matches the shape of the curve of Dhuime et al. (2012) (Fig. 2), although the latter is based on different geochemical and modelling approaches (see Section 2.1). However, it is important to note that since the Allègre and Rousseau (1984) model is about the relative volumes of crust of different ages at the present time, the $K = 15$ curve in Fig. 8A implies that ~65% of today's continental crust is >3 Ga old. Yet such large proportions of ancient crust preserved in the present continental crust remain difficult to reconcile with a number of independent observations, as discussed below.

As noted in Section 3.1, the Allègre and Rousseau (1984) growth curve(s) can be usefully compared with that of Condie and Aster (2010) (Fig. 1, curve 'C&A'), since both, in different ways, provide estimates of the variation in the proportions of juvenile rocks of different ages within the present crust. These growth models, and any other models based on this or similar approaches, make the assumption that this variation preserved in the present-day crust reflects the changes in the rates of continental growth through time (Condie, 1998; Condie and Aster, 2010; Arndt and Davaille, 2013; Arndt, 2013). The Allègre and Rousseau (1984) curve for $K = 15$ suggests that the bulk present-day crust would be made of ~65% of crust older than 3 Ga, whereas the curve of Condie and Aster (2010) (Fig. 1) implies that only ~5% of that crust is exposed at the Earth's surface. This could be accommodated, for instance, by the presence of large volumes of 'hidden' Archaean crust within the lithosphere (Begg et al., 2009; Belousova et al., 2010). However large volumes of Archaean rocks within the present crust seems unlikely when the mean age of the bulk continental crust is considered. For $K = 1$ (i.e., when no preferential erosion law is applied to the model), the model of Allègre and Rousseau (1984) generates a mean age of 1.8 Ga for the present-day crust (calculated from the relative proportions of the seven segments of ages 3.50, 2.75, 2.25, 1.75, 1.25, 0.75 and 0.25 Ga in the present crust; see Section 3.1). This age is within error of the estimate for the mean age of loess deposits of 1.82 ± 0.07 Ga recently obtained by Chauvel et al. (2014) (and see also Goldstein et al. (1984) and Goldstein and Jacobsen (1988)). The mean age of the present-day crust increases to 1.9 Ga for $K = 2$; then 2.2 Ga for $K = 4$; 2.4 Ga for $K = 6$; and 2.8 Ga for $K = 15$. Since mass balance calculations of the crust–mantle system of Allègre et al. (1983) have also constrained the mean age of the continental crust to between 2.0 and 2.4 Ga, the implication is that global values of $K \sim 15$ may be too high. It thus appears unlikely that significant volumes of old crust are hidden away somewhere at the present day, and $K = 4$ –6 may be better representative of the entire Earth system.

There is increasing evidence that continents were mostly, if not entirely, below sea level until 3.0–2.5 Ga (Arndt, 1999; Flament et al., 2008, 2011, 2013; Dhuime et al., 2015; Tang et al., 2016; Campbell and Davies, 2017; and see Section 4.2). This suggests that the preferential erosion law of Allègre and Rousseau (1984) (Fig. 7) may not have applied throughout Earth's history, and that the value of the erosion parameter K may have not remained constant through time (Cawood et al., 2013; Hawkesworth et al., 2013). Fig. 8B summarises continental growth curves calculated for a two-stage evolution of the erosion parameter K , with $K = 1$ (i.e., no preferential erosion) until 2.5 Ga and variable post-Archaeon values between 2 and 50. To avoid confusion with models that retain a uniform value of the erosion parameter K , as in Allègre and Rousseau (1984), we use the notation ' K_{2s} ' to indicate models calculated with a non-uniform, two-stage evolution of K . The differences between curves in Fig. 8A and 8B typically remain small, because the age differences between younger and older crustal sources were limited in the Archaeon, and so the correction for the preferential erosion of young rocks had relatively little effect in the estimation of the volumes of ancient crust.

3.3. Continental growth from the Nd isotope record of fine-grained sediments worldwide

The original growth model developed by Allègre and Rousseau (1984) was established from the analysis of a restricted number of sedimentary rock samples (i.e., eight shales) from the Australian continent. The question remains as to whether this dataset can be taken as representative of the crust, or whether other sets of data should be considered (Allègre and Rousseau, 1984; Michard et al., 1985; Dia et al., 1990a). To address this issue, analyses of 645 fine-grained continental sedimentary rocks sampled worldwide were compiled from the literature, and the variation of the Nd model age as a function of the deposition ages of these

sediments is plotted in Fig. 10. The variation of Nd model ages in sediments with similar depositional ages may be large, but it is striking that when the running median of the data is considered (Fig. 10, circles) the regression curve through the data has a trend very similar to that defined by the Australian shales data of Allègre and Rousseau (1984) (diamonds). It thus appears that the Australian samples can be considered as a reasonable proxy for the evolution of the continental crust on a global scale (and see also Hawkesworth et al., 2010), and that the growth curves generated by the original model of Allègre and Rousseau (1984) (Fig. 8) are not significantly affected by being based solely on the Australian shales data.

3.4. Can the Nd in fine-grained sediments data be modelled differently?

In the box-model of Allègre and Rousseau (1984) the continental crust was arbitrarily split into seven segments of ages binned at 500 Ma intervals, with the oldest segment of mean age = 3.5 Ga. This choice was based on the pioneer study of Hurley and Rand (1969), in order to simplify comparison between models. Since that time, a number of studies have provided compelling evidence for continental crust older than 3.5 Ga (e.g., Stern and Bleeker, 1998; Bowring and Williams, 1999; Wilde et al., 2001; Kemp et al., 2010), and thus for the present study a box-model with an age of 4.4 Ga for the oldest continental segment was adopted. The time-resolution of the modelling was improved by reducing the frequency of the steps from 500 Ma to 250 Ma, and the worldwide fine-grained sediments compilation (Fig. 10) was used as input data.

The discussion of how bulk rock (fine-grained sediment) and mineral (zircon) databases have been used to model the evolution of the continental crust, highlights the differences in the two approaches. The first approach (hereafter referred to as '*Model 1*') is based on the proportions of segments of different model ages in the present-day continental crust, and the second

approach (hereafter referred to as '*Model 2*') involves estimating the proportions of juvenile crust at different times in the history of the crust. Here, both approaches are applied to the worldwide compilation of Nd isotope variations in fine-grained sediments (see Fig. 10). The results are summarised in Fig. 11, and for the reasons discussed in Section 3.2, post-Archaeon values of $K_{2s} = 2, 4$ and 6 were considered. *Model 1* curves are calculated using the same method as in the original study of Allègre and Rousseau (1984). These are presented as dashed curves in Fig. 11, and they are broadly similar to the curves in Fig. 8. The calculated proportion of present-day crust >3 Ga is 14% for $K_{2s} = 2$, 21% for $K_{2s} = 4$, and 27% for $K_{2s} = 6$ (Fig. 11). The mean age of the present-day crust is 1.9 Ga if $K_{2s} = 2$, 2.0 Ga for $K_{2s} = 4$, and 2.1 Ga for $K_{2s} = 6$. *Model 2* curves (continuous curves in Fig. 11) are cumulative curves of the proportion of juvenile crust calculated at each step t_n of the box-model, using equation (6) of Section 3.1. Importantly, these curves are very similar to the zircon curve of Dhuime et al. (2012) (Fig. 11), which is reassuring in that both curves were calculated in similar ways. These curves suggest that at least ~65% of the present volume of the continental crust was established by 3 Ga, and there was a sharp change in the net rates of growth of the continental crust at that time.

These results highlight how the calculated growth curves are sensitive to the ways in which they are constructed, as in *Model 1* or *Model 2* (Fig. 11). We would argue that it is most unlikely that the present-day distribution of rocks of different model ages preserved in the continental crust (*Model 1*) is a plausible record of the volumes of crust of different ages that were present throughout Earth's evolution. We prefer approaches that rely on estimating the relative proportion of juvenile and reworked crust through time (*Model 2*), as in the growth models of Belousova et al. (2010) and Dhuime et al. (2012). Applying the *Model 2* approach to the Nd isotope variations in continental sediments reaffirms the hypothesis of Dhuime et al.

(2012) for a two-stage evolution of the continental crust based on inferred changes in the rates of continental growth. The net growth rates calculated from the different curves plotted in Fig. 11 during Stage 1 (>4 Ga to ~3 Ga) range from 2.9–3.4 km³ yr⁻¹, and they drop to 0.6–0.9 km³ yr⁻¹ on average during Stage 2 (~3 Ga to the present day).

The zircon curve of Dhuime et al. (2012) was built from a database of ~7000 detrital zircons with 'young' deposition ages ranging from the late Palaeozoic to the present day. This curve gives the lowest estimate for the volume of crust present at 3 Ga (~65%). For $K_{2s} = 1$, i.e., if no preferential erosion is applied to the Nd in continental sediments *Model 2* curves, a similar volume of 3 Ga crust is obtained. As a consequence, the slight difference between the zircon curve and the $K_{2s} = 2$ to 6 curves for continental sedimentary rocks may be explained by an overestimation of the proportion of young juvenile crust in the recent sediments from which zircons were separated. This hypothesis suggests that the zircon curve would at best give an estimate of the *minimum* volumes of crust that were established at different times, and the best approximation for the 'true' volumes of crust at these times would be given by the $K_{2s} = 4$ –6 *Model 2* curves in Fig. 11. Interestingly, *Model 2* curves for $K_{2s} = 4$ –6 give ~70–75% of the present volume of crust established by 3 Ga, which is entirely consistent with the recent estimate of Pujol et al. (2013) based on the evolution of the atmospheric ⁴⁰Ar/³⁶Ar through time (see Fig. 1, curve 'P', and Fig. 12).

4. Tectonic implications

4.1. Global onset of plate tectonics and variations in the rates of formation and destruction of the continental crust through time

The timing for the global onset of plate tectonics remains debated, and it ranges from the Hadean to the late Proterozoic depending on which proxies are considered (Kröner and Layer, 1992; de Wit, 1998; Komiya et al., 1999; Nutman et al., 2002; Stern, 2005; Brown, 2006; Cawood et al., 2006; Moyen et al., 2006; Smithies et al., 2007; Van Kranendonk et al., 2007; Dilek and Polat, 2008; Harrison et al., 2008; Hopkins et al., 2008; Pease et al., 2008; Shirey et al., 2008; Nebel-Jacobsen et al., 2010; Hamilton, 2011; Shirey and Richardson, 2011; Dhuime et al., 2012; Naeraa et al., 2012; Stern et al., 2013, 2016; Griffin et al., 2014; Ernst et al., 2016; Hastie et al., 2016); and see Fig. 12). A number of recent models of continental growth show a change in the average rate of growth at ~3 Ga (Fig. 12), and this was interpreted by Dhuime et al. (2012) as reflecting a major change in the geodynamical setting(s) in which the continental crust was formed and destroyed at that time (but see also Stern et al. (2016, 2017) for an alternate interpretation). Dhuime et al. (2012) noted that the net growth rates of $\sim 3 \text{ km}^3 \text{ yr}^{-1}$ before 3 Ga (see Fig. 11) are broadly similar to recent estimates for crust generation rates today (Scholl and von Huene, 2007, 2009; Stern and Scholl, 2010; Stern, 2011), suggesting that there may have been little or no crustal destruction at that time. In contrast, the lower net growth rates of $\sim 0.8 \text{ km}^3 \text{ yr}^{-1}$ may largely reflect higher crustal destruction rates, as subduction zones became the major locus of crust formation and destruction after 3 Ga (e.g., Cawood et al., 2006).

Shirey and Richardson (2011) compiled isotopic and bulk chemical data of silicate and sulphide inclusions in diamonds from ancient lithospheric mantle keels, and showed that mineral inclusions with eclogitic compositions were only present after ~3 Ga (Fig. 12). They

proposed that this reflects the onset of subduction and continental collision, in ways similar to that at the present day. The combination of both the diamond inclusions data and the latest continental growth models thus independently provide strong evidence for the widespread development of 'modern style' subduction zones, and the onset of the Wilson cycle of oceans opening and closing around 3 Ga. It is however important to note, as highlighted by Stern (2005) and Stern et al. (2013, 2016, 2017), that there are few occurrences of rocks and minerals that might unambiguously indicate modern-style plate tectonics processes, such as ophiolites, blueschists and glaucophane eclogites, and lawsonite eclogites (Fig. 12), older than ~1 Ga.

The differences observed between continental growth curves and age distribution curves for the juvenile crust that is preserved today (Fig. 11) can be accommodated by the destruction of large volumes of ancient crust through subduction after the onset of plate tectonics. Recent studies further suggested that ~3 Ga marked the transition between two different types of continental crust. Continental crust generated before 3 Ga was on average mafic, dense and relatively thin (<20 km) (Dhuime et al., 2015), and the upper crust sampled by sediments at that time was also mafic (Tang et al., 2016). In contrast, continental crust that formed after 3 Ga gradually became more intermediate in composition, buoyant and thicker (Dhuime et al., 2015). The rates of destruction of these two types of crust through time have been estimated at every 500 Ma (Fig. 13), using the Nd in worldwide fine-grained sediments $K_{2s} = 6$ *Model 2* growth curve as the preferred 'best estimate' for the volume of continental crust established at any time (Fig. 13, inset; and see Fig. 11, *Model 2* continuous curve, and Sections 3.2 and 3.3), and the $K_{2s} = 6$ *Model 1* curve for the proportions of continental segments of different ages which are preserved in the crust today (Fig. 13 inset; and see Fig. 11, *Model 1* dashed curve). In this preliminary model, rates of crust formation were assumed to vary smoothly through

time, with modelled mantle temperature (Korenaga, 2013), and rates decrease from ~ 4 to $3 \text{ km}^3 \text{ yr}^{-1}$ between the Hadean and the present. The model predicts that about 50% of the volume of the pre-3 Ga crust, presumably mafic and relatively dense (Dhuime et al., 2015; Tang et al., 2016), was destroyed and replaced by younger crust within the ~ 1 Ga that followed the onset of plate tectonics (i.e., from 3–2 Ga). Since ~ 2 Ga, the destruction of the post-3 Ga crust, which is arguably more differentiated, more buoyant and thicker than the pre-3 Ga crust (Dhuime et al., 2015), appears to become predominant. Finally this model predicts that at least 100% of the present volume of the continental crust has been destroyed and recycled back into the mantle since the onset of plate tectonics (Dhuime et al., *in prep.*). These estimates open new perspectives for recent models of mantle evolution and crust–mantle interaction through time (e.g., Honing and Spohn, 2016; Kumari et al., 2016; Walzer and Hendel, 2017).

4.2. ~ 3 Ga as a key transition in Earth's history

There is increasing evidence that ~ 3 Ga, and the inferred onset of plate tectonics around that time, constitute a key transition in the evolution of the Earth system (Cawood et al., 2006; Shirey and Richardson, 2011; Dhuime et al., 2012; Hawkesworth et al., 2016, 2017) (Figs. 14, 15). The secular cooling of the Earth, and associated changes in the mantle temperature and rheology, allowed the transition from a regime dominated by 'vertical tectonics' to one dominated by 'horizontal tectonics' (van Hunen et al., 2008; Sizova et al., 2010; Van Kranendonk, 2010; Korenaga, 2011, 2013; van Hunen and Moyen, 2012; Debaille et al., 2013; Gerya, 2014; Johnson et al., 2014, 2017; Gerya et al., 2015; Condie et al., 2016; Fischer and Gerya, 2016; Van Kranendonk and Kirkland, 2016; Rozel et al., 2017). The Archaean crust that is still preserved today has a bimodal silica composition (Kamber, 2015; Hawkesworth et al., 2016) and the proportion of tonalite–trondhjemite–granodiorite (TTG)

rocks that outcrops at the surface is significantly greater than that occupied by greenstone belts (Glikson, 1979; Condie, 1981; Kröner, 1985; Moyen and Martin, 2012). In contrast, the average new continental crust that was generated before 3 Ga was mafic, with Rb/Sr ~ 0.03 and SiO₂ ~ 48 –50% (Fig. 14A) and the upper crust that was present at that time had an average mafic composition (MgO $\sim 15\%$; Tang et al., 2016). This suggests that the mafic and therefore dense crust that was predominant before 3 Ga was preferentially destroyed and recycled back into the mantle after the onset of plate tectonics; whereas the TTG crust, more felsic and buoyant, remained preferentially preserved to ultimately dominate the preserved rock record.

Dhuime et al. (2015) demonstrated a systematic relationship between the Rb/Sr ratios (or the SiO₂ content) of rocks generated in modern-style subduction settings and crustal thickness at the site of sampling. They applied this relationship to estimate the evolution in thickness of the new continental crust that was formed since the onset of plate tectonics. The progressive increase in the Rb/Sr ratios of the juvenile continental crust from 3 Ga onwards (Fig. 14A) is taken to reflect the thickening of the continental crust through time, from ~ 20 km at 3 Ga to ~ 30 –40 km towards the present. In parallel, mafic rocks show a marked increase in La/Yb from ~ 2.7 Ga (Fig. 14B), along with a secular decrease in compatible element contents, and an increase in incompatible element contents, which reflect a decrease in the degrees of mantle melting through time (Keller and Schoene, 2012). These features in the mafic rocks record can be attributed to partial melting of source(s) at depths in the stability field of garnet rather than plagioclase, likely in response to thickening of the crust (Kemp et al., 2010; Kamber, 2015), the progressive development/thickening of a continental lithosphere from ~ 3 Ga onwards (Griffin et al., 2014; Hawkesworth et al., 2017), and the development of supercontinent cycles and widespread accretionary orogens after ~ 2.7 Ga (Campbell and Allen, 2008; Cawood et al., 2009) (Fig. 15).

There is a striking increase in the estimated crustal thickness (Fig. 14A) and both the rates of crustal reworking (Dhuime et al., 2012) and the global increase of the sedimentary component in the magmatic record, reflected by higher $\delta^{18}\text{O}$ in zircons, since the late Archaean (Valley et al., 2005; Spencer et al., 2014). The increase in crustal thickness is accompanied by an increasing contribution of the sediment input to the ocean, which is evidenced by increasing $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios in seawater (Veizer, 1989; Shields and Veizer, 2002; Shields, 2007), as well as increasing $\delta^{66}\text{Zn}$ in banded iron formations (BIF), a proxy for continental phosphorus input to the ocean (Pons et al., 2013), from 3 Ga onwards (Fig. 14B). These changes may be accommodated by a change in the lithosphere strength at around 3 Ga, as it became strong enough to support high-relief crust (Rey and Coltice, 2008) (Fig. 14C). This therefore indicates when significant volumes of continental crust became emergent (Arndt, 1999; Flament et al., 2008, 2011, 2013; Campbell and Davies, 2017) (Fig. 14C) and were available for erosion and surficial weathering to ultimately contribute to draw down of the CO_2 and the rise of the atmospheric oxygen at the Archaean-Proterozoic boundary (Kramers, 2002; Campbell and Allen, 2008; Kump, 2008; Lee et al., 2016).

5. Future research directions

The models outlined here predict that (i) large volumes of mafic continental crust were present early in Earth's history; and (ii) the continental crust has been generated continuously, but with a marked decrease in the rate of growth at ~ 3 Ga as subduction-driven plate tectonics started. However, it is worth considering that these results remain *model dependent* and that the debate remains open, regarding (i) the timing and the rates of the continental growth (see Fig. 1 and references therein); and (ii) when plate tectonics became a dominant process on Earth (see Fig. 12 and references therein).

There is increasing evidence that the rate of destruction of the continental crust exceeds the rate of its formation at present (Clift et al., 2009; Stern and Scholl, 2010; Stern, 2011), and that the thickness of the new continental crust generated in active margins has decreased over the last 1 Ga (see Fig. 14A). This suggests that at around 1 Ga the volume of continental crust may have been greater than it is today (Dhuime et al., 2015). This feature cannot be 'seen' with continental growth curves based on juvenile addition of new continental crust (e.g., *Model 2* curves in Fig. 11), nor on the age distribution of rocks of different ages in the crust (e.g., *Model 1* curves in Fig. 11), because such cumulative curves do not allow crustal volumes to decrease over time. This highlights the need in developing continental growth models in which the volume of continental crust may have exceeded its present value in the past (e.g., Fyfe, 1978; Stern and Scholl, 2010; Hawkesworth et al., 2016). Such models have to take into account the timing and rates at which continental crust is destroyed and recycled back into the mantle, as well as the effects of recycling process, to account for variations in the mantle chemistry both at present and back in time (e.g., Allègre, 1982; McCulloch and Bennett, 1994; Nägler and Kramers, 1998; Kramers, 2002; Xie and Tackley, 2004; Delavault et al., 2016a; Kumari et al., 2016).

'Traditional' radiogenic isotope systems (i.e., Sm-Nd, Lu-Hf, U-Th-Pb, Rb-Sr) have proven very useful in crust–mantle interaction studies (e.g., Allègre, 2008). However, uncertainty over the isotope composition of the ambient mantle in the Hadean/Archaean remains problematic for precise mass balance calculations and the determination of model ages of juvenile crust formation with these systems. This highlights the need in exploring the information provided by 'alternate' radiogenic isotope systems. The K-Ca system, with a half-life of ~1.3 Ga, is particularly suited for this purpose because the (low K/Ca) mantle shows

little variation in its $^{40}\text{Ca}/^{44}\text{Ca}$ ratio through time, whereas the (high K/Ca) continental crust has developed highly radiogenic $^{40}\text{Ca}/^{44}\text{Ca}$ ratios in the early stages of Earth's evolution (Kreissig and Elliott, 2005). Finally the impact of the 4.1–3.8 Ga late heavy bombardment (LHB) on the destruction and the recycling of the Hadean crust (Marchi et al., 2014; Shibaike et al., 2016), and on the composition of the mantle, needs to be carefully considered when developing more realistic models for the continental growth.

Although there is a consensus that subduction operated in the Archaean, but possibly only locally and in a fashion very different as today, the timing for the transition between initial 'stagnant lid' tectonic regime, during which the Earth was dominated by vertical tectonics, to a 'modern plate tectonics' regime with widespread seafloor spreading centres and deep subduction zones, remains poorly constrained (see Fig. 12). Also importantly, whether this transition was sharp (10's Ma) or progressive (100's to 1000's Ma), and whether it was synchronous on a global scale yet has to be determined (see Fig. 15). The trace element contents of bulk rocks have been used for decades as a proxy for the geodynamical context of magmas generation (e.g., Pearce and Cann, 1973). However, their use in ancient (e.g., Archaean) mafic or felsic rocks as plate tectonics indicators has remained inconclusive (e.g., van Hunen and Moyen, 2012; Condie, 2015). New perspectives in bulk-rock analysis arise from the recent development of so-called 'non traditional' isotopes such as Ti (Millet et al., 2016) and Mo (Freymuth et al., 2015, 2016), which fractionate differently in magmas generated in subduction or intraplate setting. The analysis of Ti and Mo isotopes in Archaean samples can be seen as an interesting perspective, although the behaviour of these isotopes in ancient and potentially altered/metamorphosed rocks yet has to be tested.

The *in situ* measurement of trace elements in zircon by ion probe or laser ablation techniques has a high potential for unravelling the nature and the geodynamical context of formation of the granitic magmas from which zircons crystallise (Belousova et al., 2002; Wang et al., 2012; Yang et al., 2014; Grimes et al., 2015; Trail et al., 2015; Gao et al., 2016; Smythe and Brennan, 2016; Burnham and Berry, 2017). Since the large majority of zircons (80–90%, see Section 1) crystallise from non-juvenile sources, the trace element contents in zircon may not indicate the geodynamical context of the formation of the new continental crust. However, there is potential in developing studies in which trace elements in well-dated zircons are used as a proxy for the timing, rates and the geodynamical conditions of crustal reworking through time.

Within the last few years, there has been a growing interest in developing studies based on the mineral inclusion record in zircon. Initial studies established a link between the mineralogy of the inclusions and the composition of the parent magma of the host zircon (Maas et al., 1992; Cavosie et al., 2004; Hopkins et al., 2008, 2010; Bell et al., 2011), and between the type and the chemistry of the inclusions and the presence of event(s) subsequent to the crystallisation of the host zircon (Bell et al., 2015b; Bell, 2016). More recently Jennings et al. (2011) and Bruand et al. (2016, 2017) established a strong correlation between the trace element content of apatite, both within the rock matrix and in inclusion within zircon, and the composition of the bulk rock. In parallel, Dhuime et al. (2014) and Delavault et al. (2016b) explored the potential of combining *in situ* isotopic analyses of zircons and their mineral inclusions to calculate the time-integrated Rb/Sr and U/Pb ratios of the juvenile continental crust at the time of its Hf model age. The time-integrated Rb/Sr ratio can be used as a proxy for the composition and the thickness of the juvenile continental crust (Dhuime et al., 2015, and see Fig. 14A). It is calculated from the Sr isotope composition of apatite inclusions and the U-Pb

ages and the Hf isotope ratios of the host zircon (Dhuime et al., 2014). The time-integrated U/Pb ratio can be used to unravel the tectonic setting of formation of the juvenile continental crust (Delavault et al., 2016b). It is calculated from the Pb isotope composition of feldspar inclusions and the U-Pb ages and the Hf isotope ratios of the host zircon (Delavault et al., 2016b). Studies based on the analysis of inclusions in large collections of zircons of both magmatic and detrital origin therefore have significant potential for unravelling the composition and the tectonic setting(s) of formation of the new continental crust through time, in ways that have never been explored before.

Finally, the recent development of increasingly more sophisticated numerical codes has allowed models for the dynamics of crust–mantle interaction to be developed in ways that were difficult to achieve until now (van Hunen and Moyen, 2012; Korenaga, 2013; Johnson et al., 2014, 2017; Gerya et al., 2015; Fischer and Gerya, 2016; Honing and Spohn, 2016; Rozel et al., 2017; Walzer and Hendel, 2017). The integration in such models of the expanding large geochemical databases, including those available through online platforms (e.g., GEOROC: <http://georoc.mpch-mainz.gwdg.de/georoc/>; EarthChem: <http://www.earthchem.org/portal>), should give the opportunity to initiate more accurate global models that take account of the inter-relationship and the linked variations between plate tectonics, magmatic activity, continental growth, the oxygenation of the atmosphere, and the development and the evolution of life on Earth (e.g., Arndt and Nisbet, 2012; Philippot et al., 2012; Cawood and Hawkesworth, 2014; Lee et al., 2016; Stern, 2016; Duncan and Dasgupta, 2017).

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Figures caption

Fig. 1. Models for the volume of the continental crust through time, taken from the literature data. F: Fyfe (1978); Br: Brown (1979); O’N: O’Nions et al. (1979); V&J: Veizer and Jansen (1979); A: Armstrong (1981); D&W: Dewey and Windley (1981); Al: Allègre (1982); M&T: McLennan and Taylor (1982); A&R: Allègre and Rousseau (1984); R&S: Reymer and Schubert (1984); P&A: Patchett and Arndt (1986); M&B: McCulloch and Bennett (1994); T&M: Taylor and McLennan (1995); K&T: Kramers and Tolstikhin (1997); C&K: Collerson and Kamber (1999); C: Campbell (2003); R: Rino et al. (2004); Be: Belousova et al. (2010); C&A: Condie and Aster (2010); D: Dhuime et al. (2012); P: average curve of Pujol et al. (2013).

Fig. 2. Continental growth models based on the combination of Hf and U-Pb data from worldwide sets of zircon data. The black curve (Belousova et al., 2010) is calculated from a compilation of 13,844 zircons mostly of detrital origin, and the blue curve (Dhuime et al., 2012) is calculated from a compilation of 6972 zircons from young sediments, in which the contribution of so called 'hybrid' Hf model ages was corrected from the worldwide Hf-O relationship evidenced by Dhuime et al. (2012) (inset).

Fig. 3. Influence of using different inferred reservoirs from which new continental crust was extracted in (A) the calculation of crust formation ages, and (B-C) continental growth curves. References for these reservoirs are given in the text. Growth curves calculated from the depleted mantle (green curve in panel A) and an alternate reservoir (orange curve) are shown in panels B (green curve) and C (orange curve), respectively. The growth curve of Dhuime et al. (2012), calculated from the new crust reservoir (black curve in panel A), is shown in panels B-C for comparison (grey thick curve).

Fig. 4. Influence of using different inferred $^{176}\text{Lu}/^{177}\text{Hf}$ ratios for the crustal source of the magma from which zircons crystallised in (A) the calculation of crust formation ages, and (B-C) continental growth curves. Growth curves calculated with $^{176}\text{Lu}/^{177}\text{Hf} = 0.009$ (red evolution path in panel A) and $^{176}\text{Lu}/^{177}\text{Hf} = 0.022$ (green evolution path) are shown in panels B (red curve) and C (green curve), respectively. The growth curve of Dhuime et al. (2012), calculated with $^{176}\text{Lu}/^{177}\text{Hf} = 0.015$ (grey evolution path in panel A), is shown in panels B-C for comparison (grey thick curve).

Fig. 5. Influence of ancient radiogenic Pb loss on (A) the calculation of model ages, and (B) the shape of the continental growth curve calculated with the equations and the database of Dhuime et al. (2012), in which ancient Pb losses were simulated using an in-house Excel spreadsheet. The original growth curve of Dhuime et al. (2012) is shown in panel B for comparison (grey thick curve).

Fig. 6. Influence of using databases composed of zircons of both detrital and magmatic origin, from rocks with a range of deposition and crystallisation ages, on the shape the continental growth curve calculated with the equations of Dhuime et al. (2012). (A) Belousova et al. (2010) database ($n = 13,832$); (B) Roberts and Spencer (2015) database ($n = 42,469$). The growth curve calculated from the Dhuime et al. (2012) database ($n = 6972$) is shown in panels B-C for comparison (grey thick curve).

Fig. 7. Schematic representation of the two-block model of Allège and Rousseau (1984), which they used to evaluate the bias induced in the sedimentary record by the preferential erosion of young (juvenile) high-relief crust over older lower relief crust. This bias is

expressed through and erosion parameter K that relates the variation in the relative proportions of the source rocks in the sediment to those within the crust.

Fig. 8. Continental growth curves calculated using the model, the equations and the data of Allègre and Rousseau (1984). The shape of these curves critically depends on the value of the erosion parameter K used in the equations, and values ranging between 2 and 50 were chosen for comparative purposes. (A) Continental growth curves calculated with values of K constant throughout Earth's history. (B) Continental growth curves calculated for a two-stage evolution of K , with $K = 1$ (i.e., no preferential erosion) until 2.5 Ga and variable post-Archaeon values between 2 and 50. The notation ' K_{2s} ' is used for this 2-stage erosion parameter.

Fig. 9. Profile of the Frankland River, southwestern Australia, after Cawood et al. (2003), including sediment sample locations (stars) and boundary between Yilgarn and Albany-Fraser crustal segments (dashed vertical line). The values of the erosion parameter K calculated by Dhuime et al. (2011b) are reported for each sample.

Fig. 10. Nd model age as a function of the age of deposition of worldwide fine-grained continental sedimentary rocks compiled from the literature data (Hamilton et al., 1983; O'Nions et al., 1983; Allègre and Rousseau, 1984; Frost and O'Nions, 1984; Miller and O'Nions, 1984, 1985; Davies et al., 1985; Hensel et al., 1985; Michard et al., 1985; Andre et al., 1986; Miller et al., 1986; Frost and Winston, 1987; Barovich et al., 1989; Mearns et al., 1989; Dia et al., 1990a, 1990b; Yanez et al., 1991; Boher et al., 1992; Nögler et al., 1992, 1995; Stevenson and Turek, 1992; Alibert and McCulloch, 1993; Turner et al., 1993; Bock et al., 1994; Jahn and Condie, 1995; Stevenson, 1995; Cullers et al., 1997; Henry et al., 2000; McLennan et al., 2000; Ugidos et al., 2003; Krogstad et al., 2004; Armendariz et al., 2008;

Lopez-Guijarro et al., 2008; Yu et al., 2009; Cabral et al., 2013). The running median of the data (circles) was calculated for every 100 Ma step, and a regression curve was calculated (brown curve, with $R^2 = 0.96$). The Australian shales data used in the pioneer study of Allègre and Rousseau (1984) (diamonds) are plotted for comparison.

Fig. 11. The volume of continental crust as a function of time, calculated using the variation in Nd isotope ratios in fine-grained sediments data plotted in Fig. 10, and (A) the original box-model and equations of Allègre and Rousseau (1984) (*Model 1* dashed curves), (B) the model developed in this study (*Model 2* continuous curves). Values of the two-stage erosion parameter K_{2s} , with post-Archaeon values of 2, 4 and 6, were chosen for the calculations. The continental growth curve of Dhuime et al. (2012) calculated from detrital zircons data is plotted for comparison.

Fig. 12. A summary of continental growth curves calculated from Nd in worldwide fine-grained sediments data (this study, *Model 2* curves in Fig. 11), along with models obtained with different geochemical approaches and recently published by Dhuime et al. (2012) and Pujol et al. (2013). The Os isotope data of sulphide inclusions within diamonds from the subcontinental mantle lithosphere are plotted on the secondary Y-axis for comparison (Shirey and Richardson, 2011). Eclogitic diamonds (pink diamonds) are distinguished by $^{187}\text{Os}/^{188}\text{Os}$ ratios higher than those of peridotitic diamonds (green diamonds) that plot along the mantle evolution curve (black curve). The green, blue and purple histograms on the top of the figure show the distribution of the 'plate tectonic and subduction indicators' of Stern et al. (2013, 2016). The vertical arrows indicate estimates for the timing of the onset of plate tectonics, taken from on a number of models from the literature data. References are indicated by the numbers in square brackets ([1]: Harrison et al. (2008); [2]: Hopkins et al. (2008); [3]: de Wit

(1998); [4]: Shirey et al. (2008); [5]: Komiya (1999), Dilek and Polat (2008); [6]: Nutman et al. (2002); [7]: Kroner and Layer (1992); Griffin et al. (2014); [8]: Moyen et al. (2006), Van Kranendonk et al. (2007), Pease et al. (2008), Naeraa et al. (2012); [9]: Smithies et al. (2007); [10]: Cawood et al. (2006); [11]: Shirey and Richardson (2011); Dhuime et al. (2012); [12]: Brown (2006); [13]: Nebel-Jacobsen et al. (2010); [14]: Stern (2005); [15]: Hamilton (2011)).

Fig. 13. Model for the evolution of the rates of destruction of the continental crust through time. Two types of crust are considered: pre-plate tectonics crust that was generated before 3 Ga, and subduction-related crust that was formed from 3 Ga until the present day. The rates of crust formation were assumed to decrease gradually from $4\text{--}3\text{ km}^3\text{ yr}^{-1}$ between the Hadean and the present. The *Model 2* $K_{2s} = 6$ growth curve presented in Fig. 11 was used as the best estimate for the volumes of continental crust that were established at any time (blue curve, inset), and the rates of crust destruction were adjusted at every 500 Ma in order to accommodate the variations in the proportions of juvenile continental segments of different ages preserved at present (*Model 1* brown curve for $K_{2s} = 6$, inset)

Fig. 14. A summary of key changes that occurred throughout the history of the Earth, with an emphasis for the period around 3 Ga. (A) Variations in Rb/Sr, SiO₂ and thickness of the juvenile continental crust (Dhuime et al., 2015), and in the La/Yb in mafic rocks (Keller and Schoene, 2012) (secondary green Y-axis and green symbols). (B) Variations in the Zn isotope composition in banded iron formations (BIFs) (Pons et al., 2013), and in the influence of the river runoff versus the mantle influence in the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in seawater (Shields, 2007) (secondary blue Y-axis and blue curve). IRCS: igneous rocks and clastic sediments average zinc isotope composition. (C) Topography (plateau elevation) that continents were able to sustain depending on the variation in the rheology of the lithosphere through time (Rey and

Coltice, 2008), and the percentage of land surface (relative to the present Earth surface) that was emerged through time (secondary brown Y-axis and brown envelope curve) (Flament et al., 2013).

Fig. 15. Schematic representation of a crust–mantle cross-section of the Earth, before and after the onset of plate tectonics around 3 Ga.

Figure 1 - 1.5 column

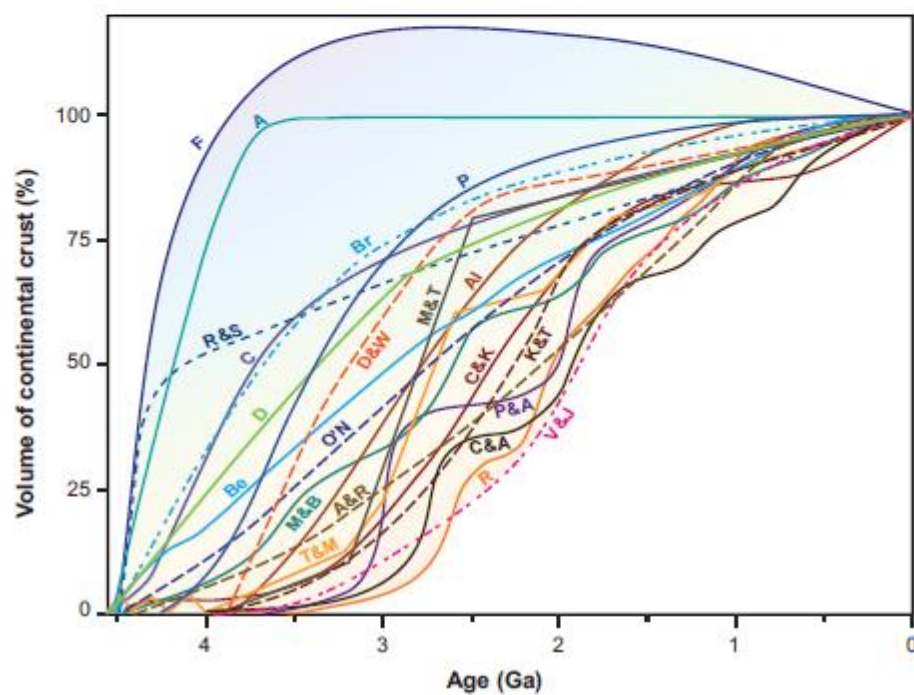


Figure 1

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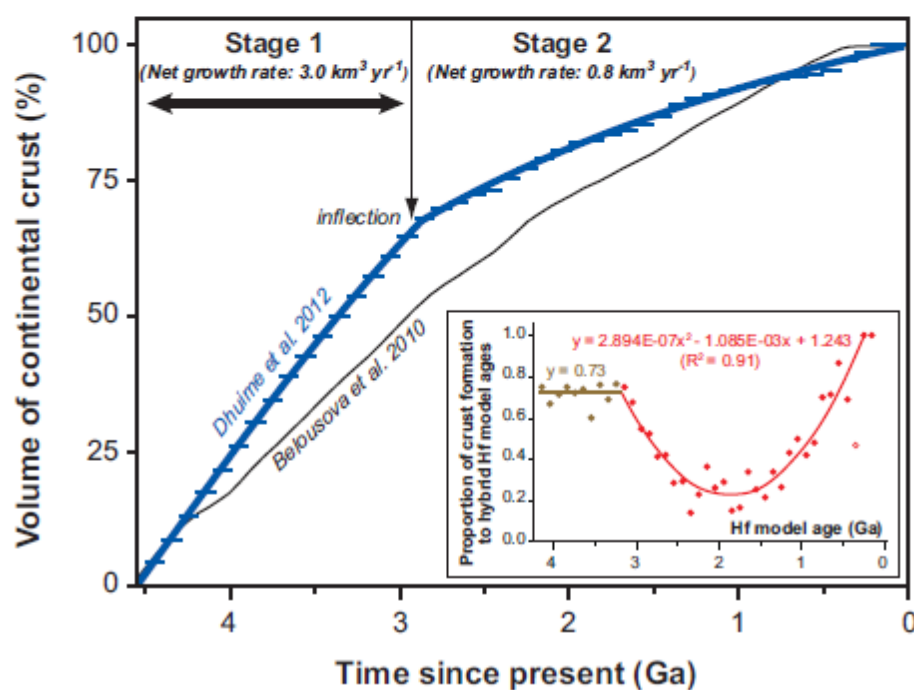


Figure 2

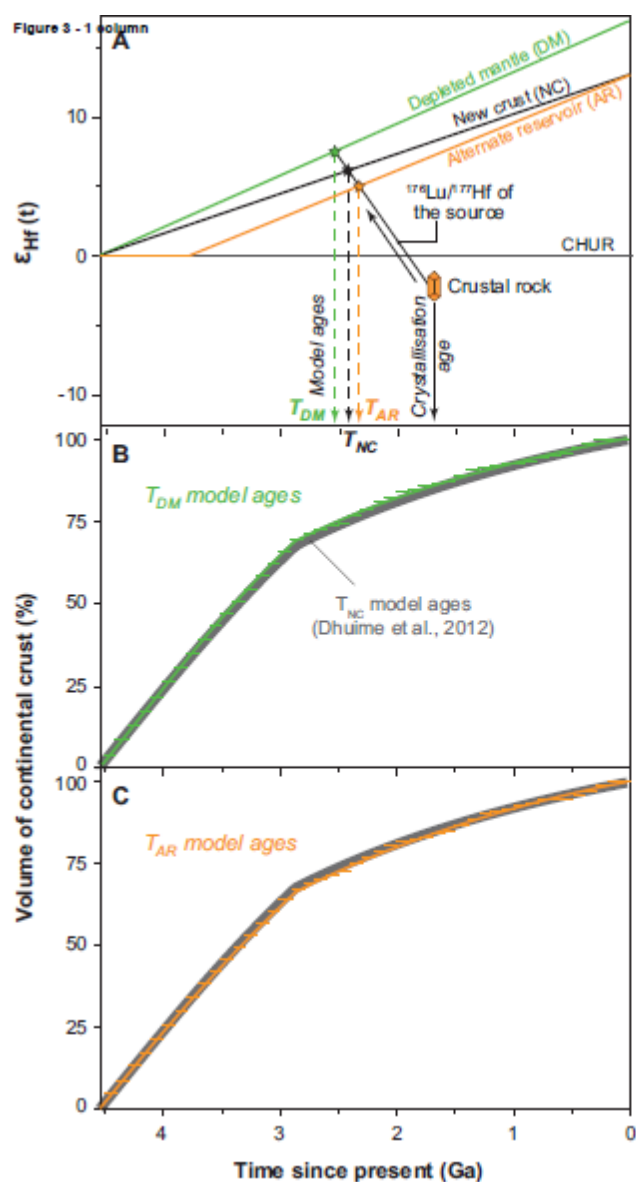


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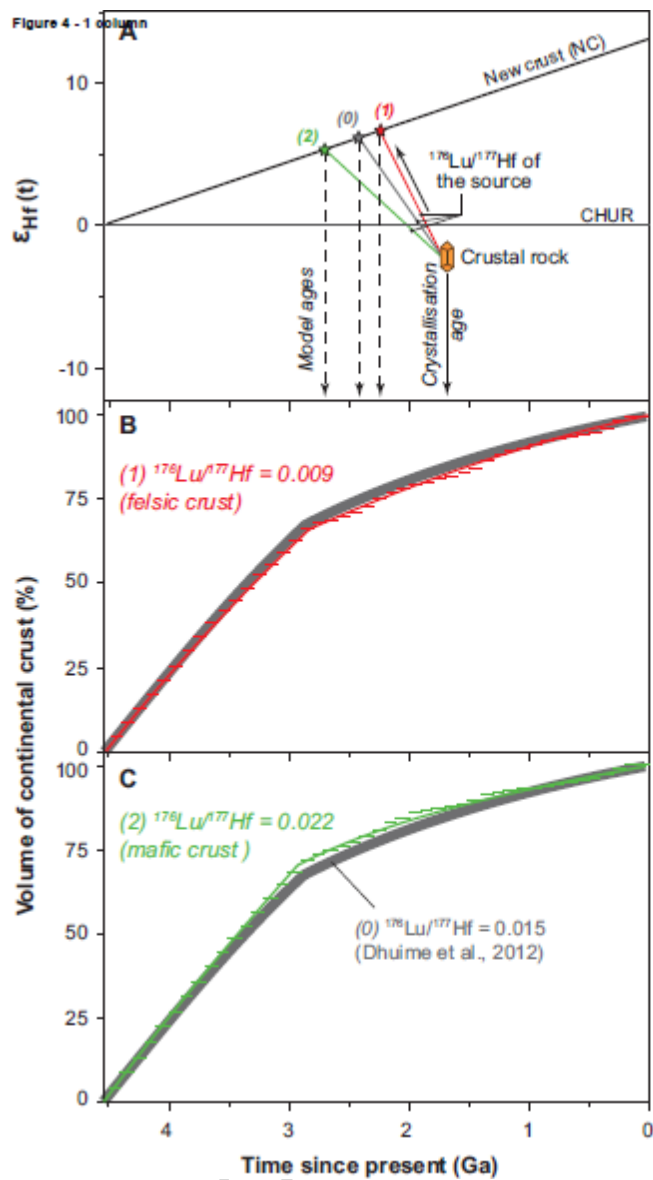


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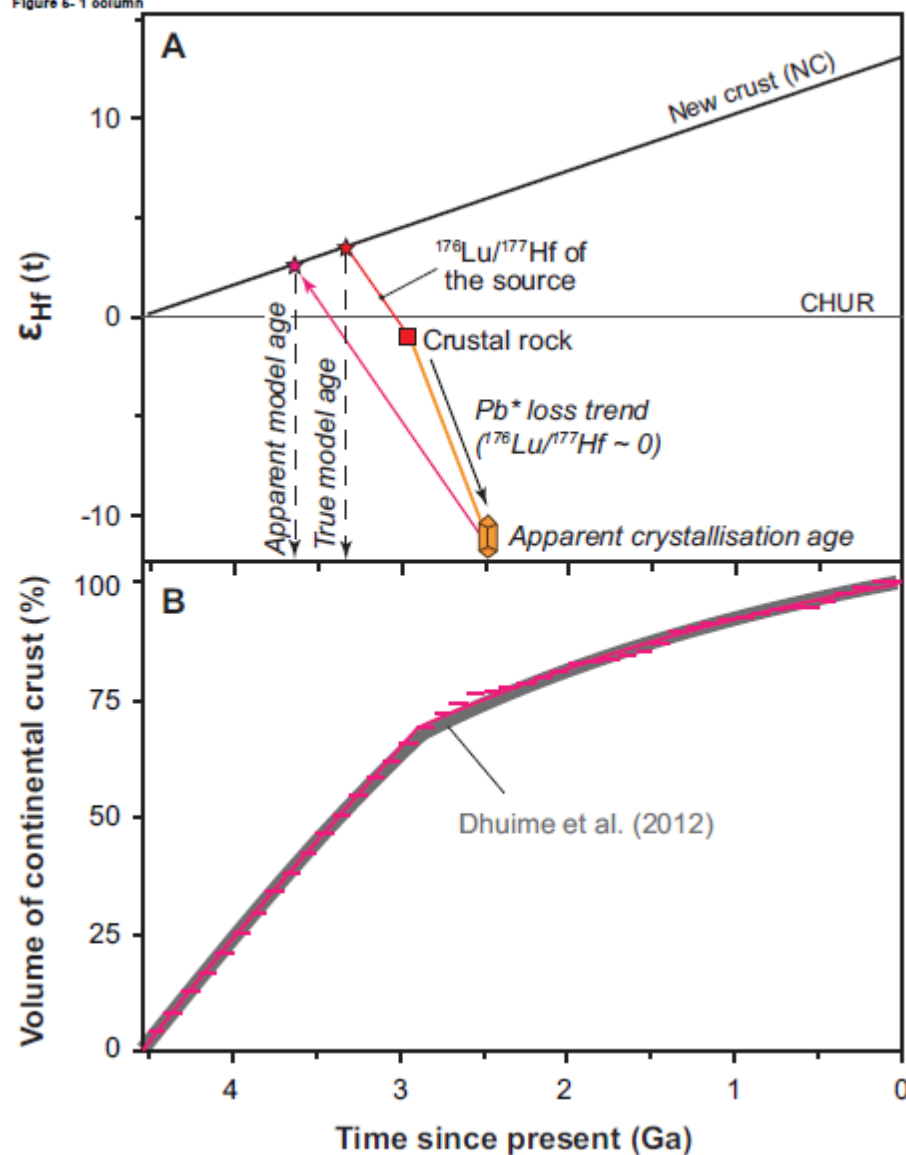


Figure 5

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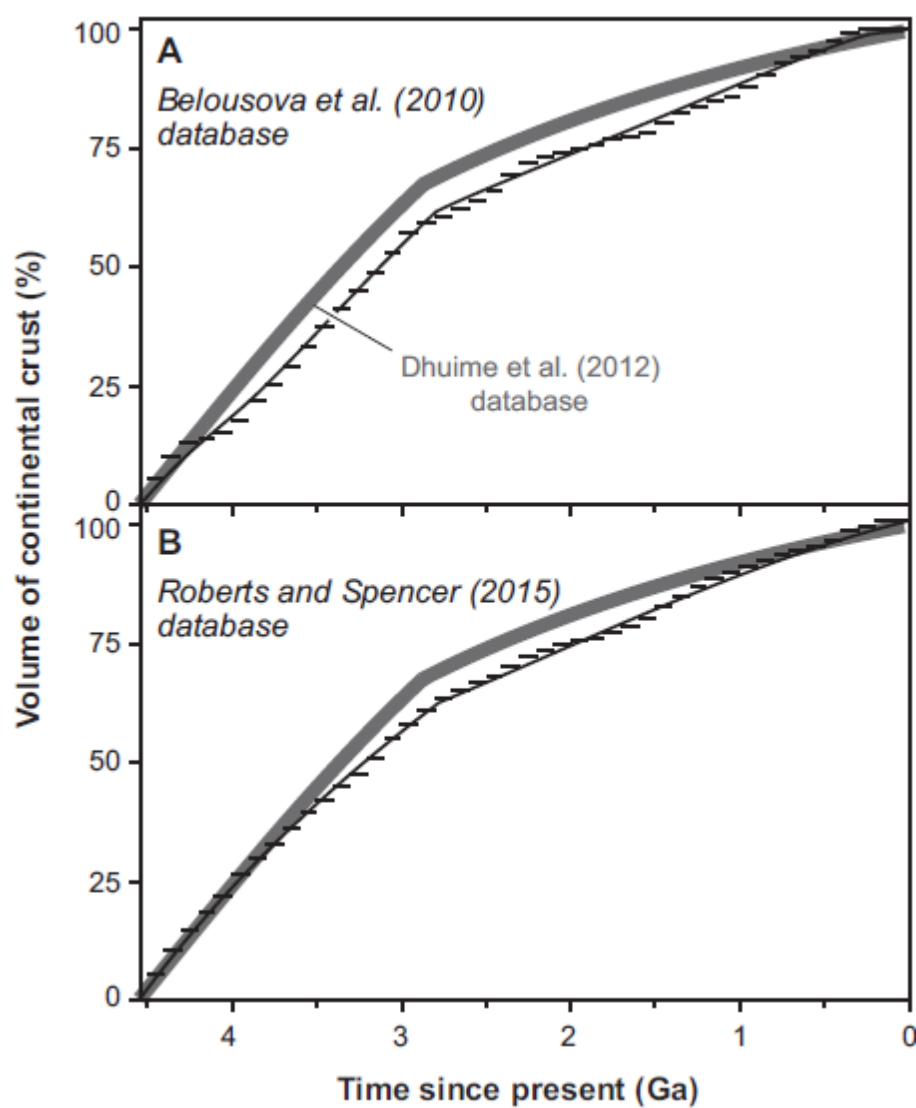


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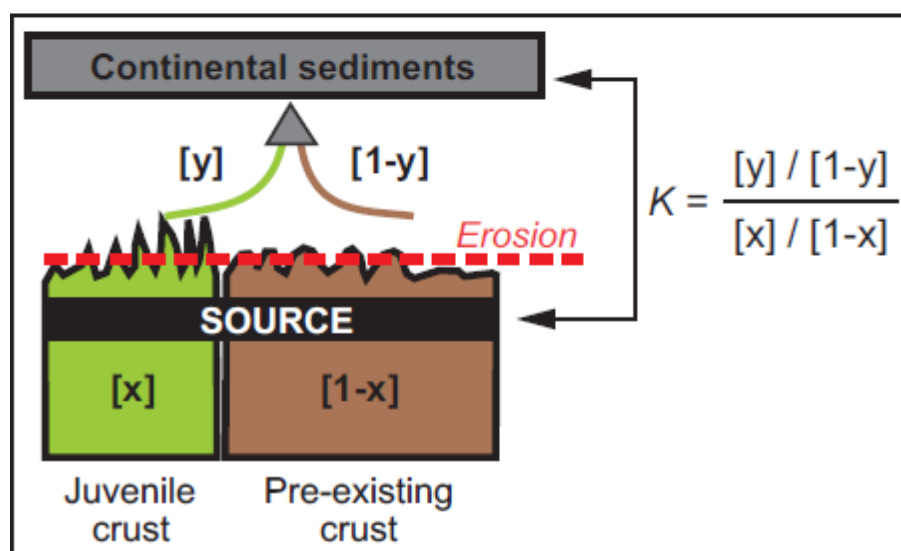


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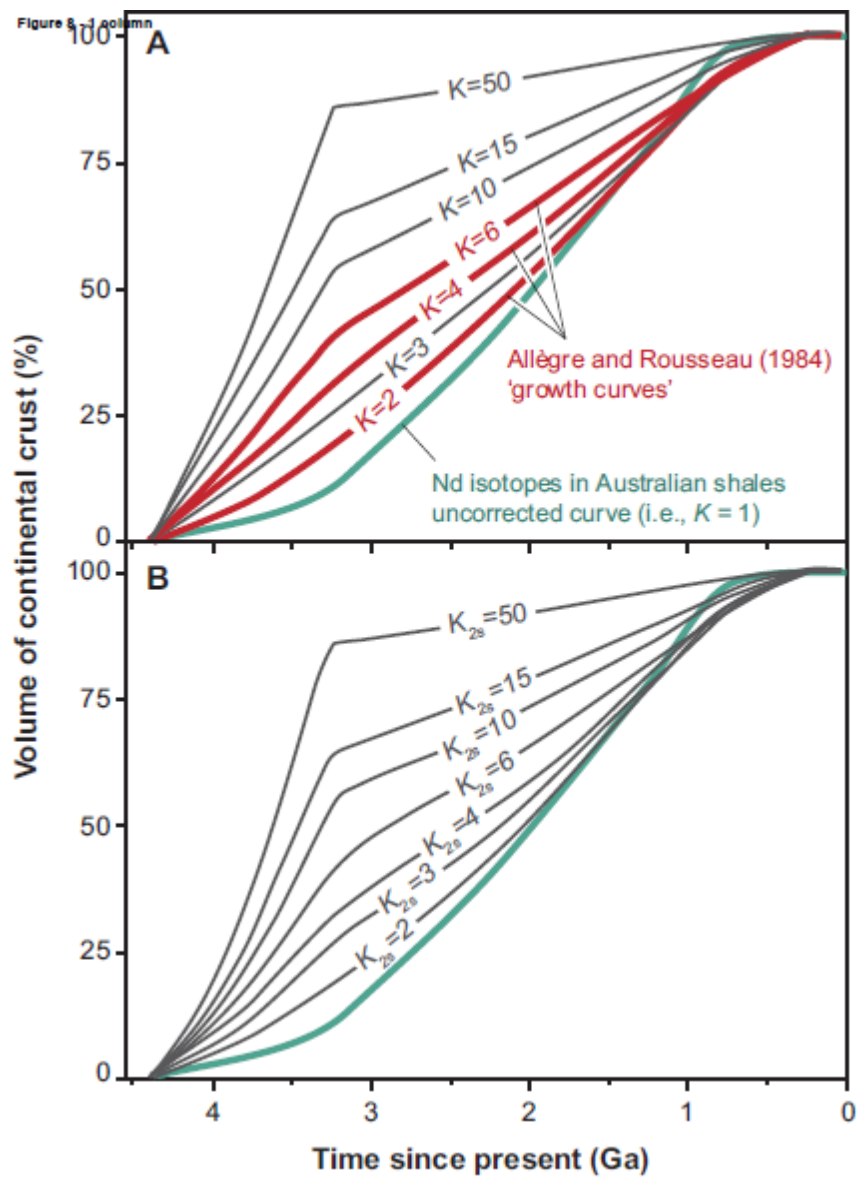


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Figure 9 - 1 column

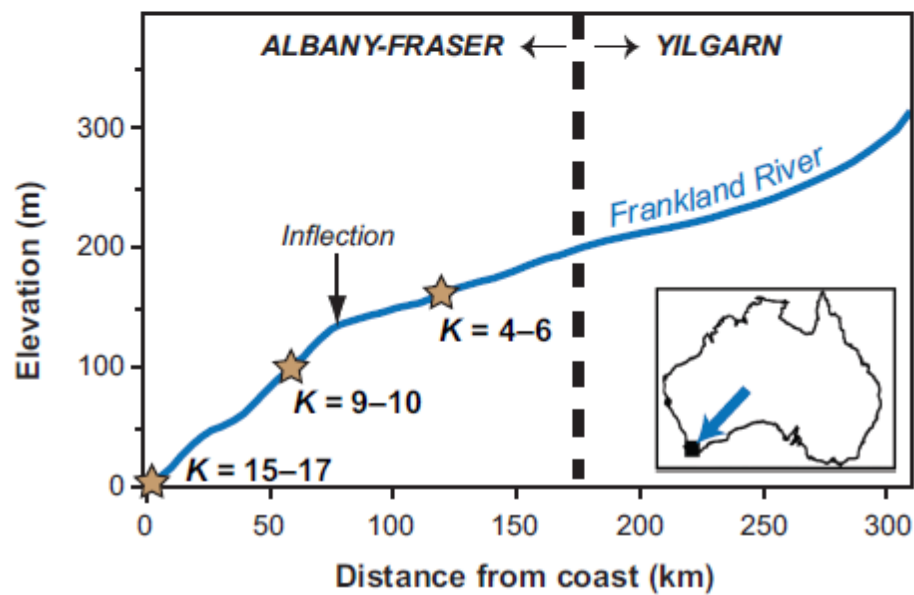


Figure 9

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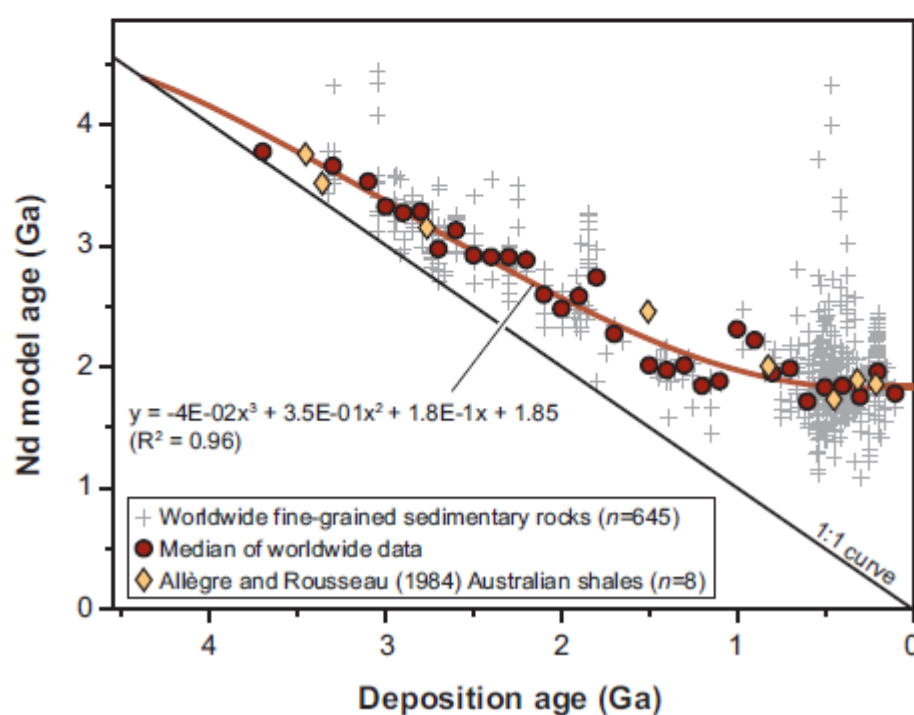


Figure 10

Figure 11 - 1 column

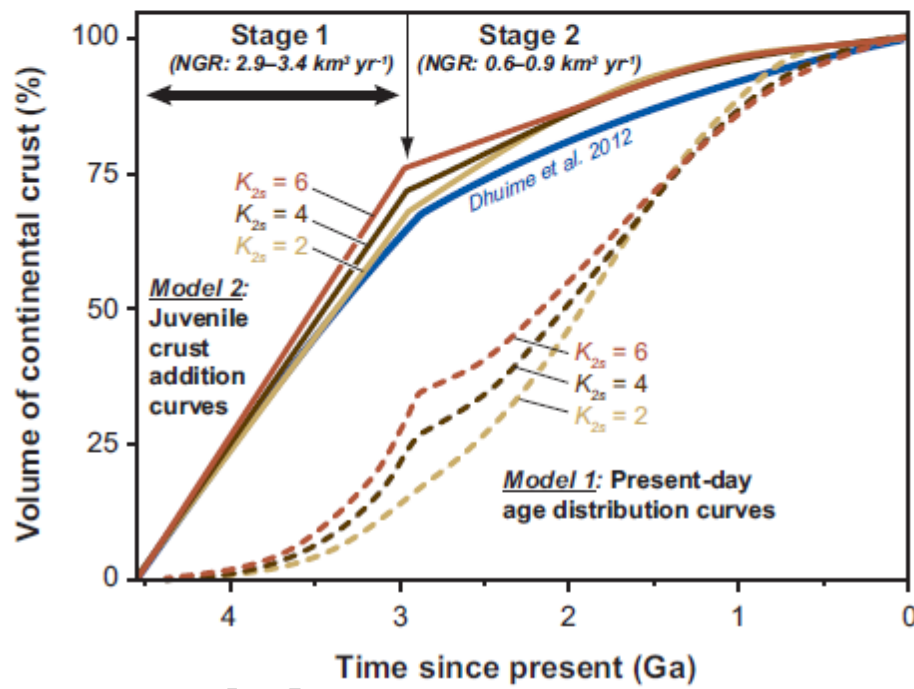


Figure 11

Figure 12 - 1.6 column

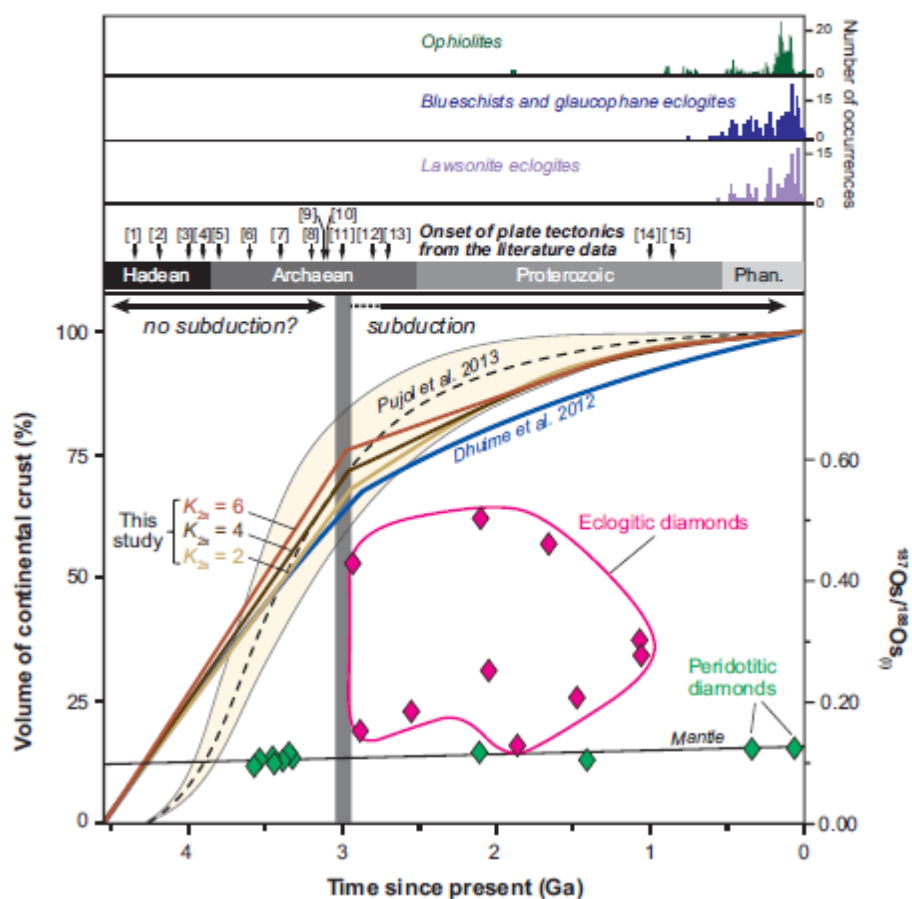


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Figure 13 - 1 column

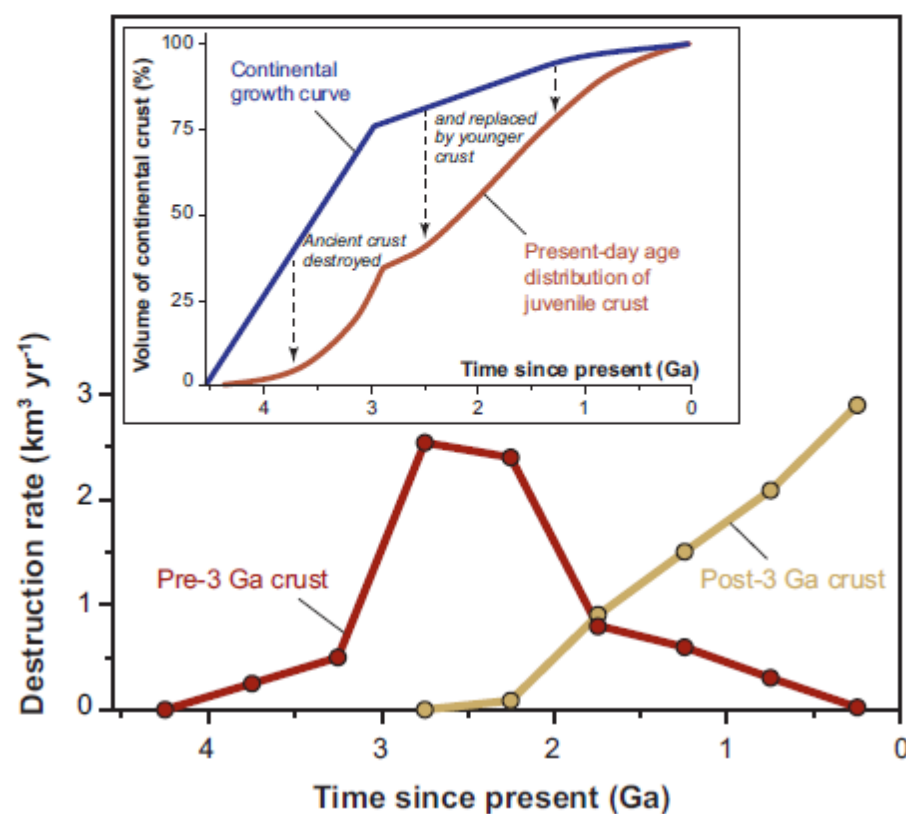


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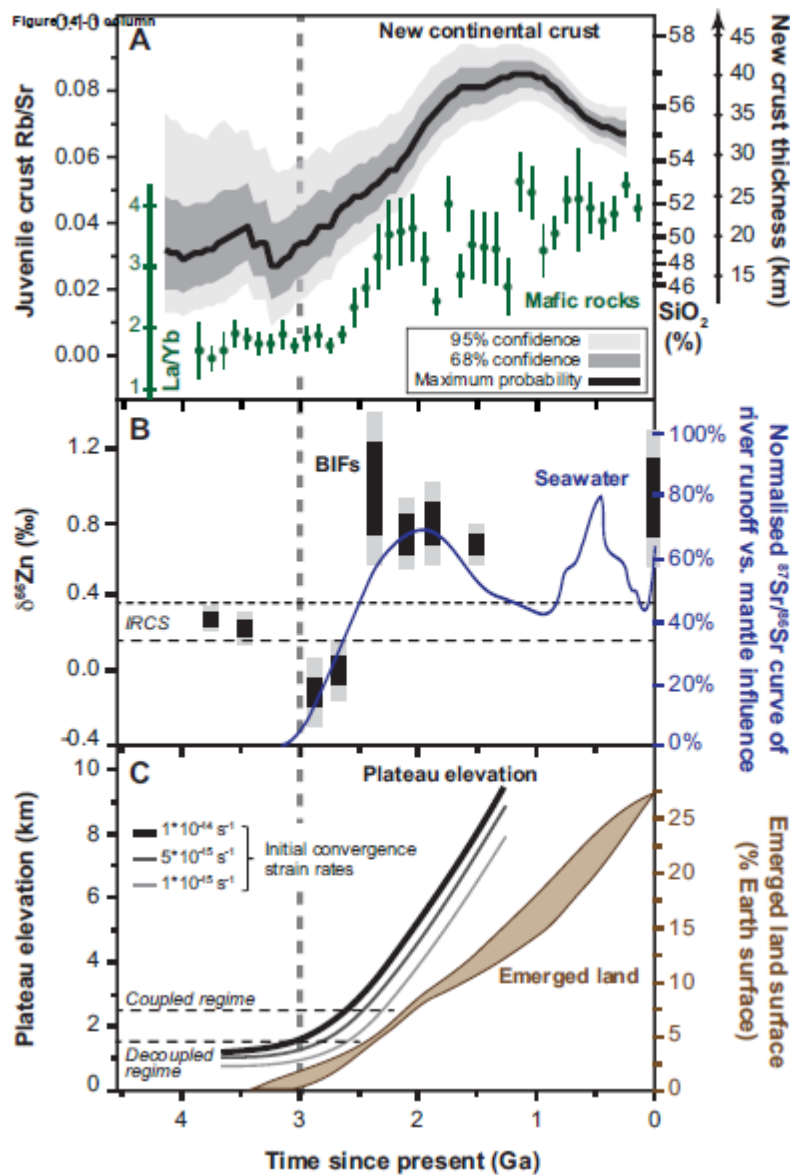


Figure 14

Figure 15 - 2 columns

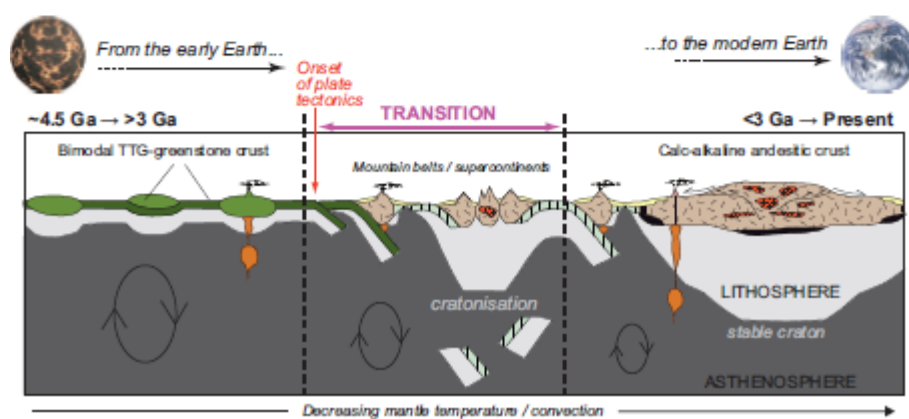


Figure 15